The Retrieval of the Effective Radius of Snow Grains and Control of Snow Pollution with GLI Data

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INTRODUCTION

The core of snow remote sensing is the connection between microphysical and radiative characteristics of snow layer. The conventional way to establish the relation between radiative and retrieved microphysics characteristics is the use of the Mie theory to obtain the single-scattering characteristics in terms of microphysics ones and the radiative transfer theory to relate radiative properties of snow layers to the single-scattering (inherent) characteristics.

It is well known that a snow layer is a multiple scattering close packed medium with irregular shaped nonunifom grains. Despite this, up to now all developed algorithms to retrieve the snow effective grain size from measured spectral reflection characteristics (Greenfell et al. (1994), Stamnes et al. (1997)) employ simple snow model as a layer with large independent spherical scatterers. This model was developed in the classical works of C.F. Bohren, S.G. Warren and W.J. Wiscombe (Bohren, (1974) , Wiscombe and Warren (1980) , Wiscombe and Warren (1980a)).

In this work we present the new algorithm to retrieve the effective snow grain size and pollution amount, which employs no *a priori* suggestion of snow grain shapes. The algorithm uses the multispectral information provided by a satellite optical instrument, particularly, by GLI.

In so doing, , we use the snow model as non-spherical close-packed grains instead of the common model of independent spherical scatterers; the geometrical optics (GO) approximation instead of the diffraction theory; the asymptotic solution instead of the radiative transfer equation.

EFFECT OF THE CHOSEN SNOW MODEL ON THE RETRIEVED EFFECTIVE GRAIN SIZE IN THE "FITTING" APPROACHES.

The input of the satellite retrieval algorithm is the reflection function $R(\theta,\theta_0,\varphi,\lambda)$ for the pixel of interest. Here θ_0 is the Sun polar angle, θ, φ are polar and azimuth angles of an observed pixel. Fig.1 shows that reflection functions for models with different particle shapes differ drastically.

The main idea of the common retrieval algorithm is the following. The data set for the radiance coefficients of the light reflected by snow with different grain sizes is computed preliminary using some radiative transfer code and

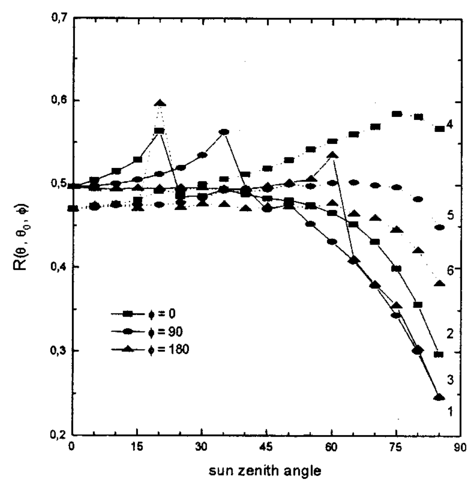

Fig. 1 Radiance coefficient as a function of the sun zenith angle at the observation angle 20° and different azimuth angles for large spherical particles (1-3) and fractals (4-6) at single scattering albedo 0.984.

the Mie theory. The a_{ef} value is retrieved as that, which provides the best fitting of measured and computed radiance coefficients ($a_{ef} = 3 < V > / < \Sigma >$ is the effective radius of grains, $< V >$ and $< \Sigma >$, the average volume and surface area of snow grains respectively). The difference between a model used in computation and real snow may generate considerable error of the retrieved value, which depends on the observation geometry. Our simulations have shown that the retrieved snow grain size ranges from $0.1\,a_{ef}$ up to a_{ef}, where a_{ef} is the true size, when using a model with spherical particles while real snow contains fractals.

Thus, because of practically nothing is known about snow grain shape in particular snow under observation, the use of any *a priori* chosen model can lead to the non-predicted and sometimes very large errors.

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EQUATIONS TO RETRIEVE THE SNOW GRAIN SIZE

The basic equations of our algorithm is the asymptotic formula for the reflection function (Zege et al. (1991)):

$$R_t(\theta,\theta_0,\varphi) = R_0(\theta,\theta_0,\varphi)\times \\ \times \exp\left\{-4y\frac{u(\theta)u(\theta_0)}{R_0(\theta,\theta_0,\varphi)}\right\}. \qquad (1)$$

Here $y=\sqrt{\dfrac{\sigma_{abs}}{3\sigma_{ext}(1-g)}}$, $R_0(\theta,\theta_0,\varphi)$ is the reflection function for the semi-infinite layer with the same phase function with no absorption ($\sigma_{abs}=0$), g, the asymmetry parameter of the phase function. σ_{ext}, σ_{sca}, σ_{abs}, the extinction, scattering and absorption cross sections, respectively. $u(\theta)=\frac{3}{7}[1+2\cos(\theta)]$ is the Miln's problem solution for such non absorbing medium. Functions $R_0(\theta,\theta_0,\varphi)$ and $u(\theta)$ do not depend on the absorption. The only value, which depends on the azimuth angle, is the function $R_0(\theta,\theta_0,\varphi)$. It is also the only function in Eq.1, which is essentially affected by shape of grains.

Using the large body of reflection functions computed with an accurate radiative transfer code, it was shown, that errors of Eq.(1) are less than 2% over the spectral range where the snow albedo is more than 0.3 (this range is used in our algorithm).

ON CLOSE-PACKING EFFECT

As known, the concentration of snow grains in snow fields is around 30-40%. It means, that the packing effect should be taken into account. It is a well-known fact that close packing effects reduce scattering cross section σ_{sca} and values of the phase function at small angles (Ivanov et al., 1988). Analysis of the experimental data has showed that the effects of close packed media are negligible for the value of σ_{abs} in the case of weak absorbing large grains. The geometrical optics (GO) part of the phase function does not depend on the volume concentration effects (Ivanov et al., 1988).

So, for the value of the transport extinction coefficient $\sigma_{ext}(1-g)$, which included in Eq.(1), we can obtain the following equation for weak absorbing grains:

$$\sigma_{ext}(1-g)=\sigma_{sca}^{GO}(1-g^{GO}), \qquad (2)$$

here, σ_{sca}^{GO}, the geometrical optics scattering cross sections, g^{GO}, the asymmetry parameter of the geometrical optics part of the phase function.

This equation can be obtained from the formulas for rare media, when close packing effects are neglected. Thus we underline that for snow close packing effects do not change the transport extinction coefficient $\sigma_{ext}(1-g)$ and the similarity parameter y. As it could be shown, the basic relation (1) follows directly from common symmetry properties of a semi-infinite weak absorbing layer independently of the radiative theory. It means that for snow in visible and near IR the basic relation (1) are valid for any value of the grain packing coefficient.

In the range 0.3 ÷ 1.25 μm, where wave attenuation along the particle size is small, the following equation for parameter y has been found:

$$y=\frac{1}{4}A\sqrt{\alpha a_{ef}}, \qquad (3)$$

Here $\alpha=4\pi\chi/\lambda$, χ, the imagine part of refractive index.

The calculation shows that for spheroidal, spherical and fractal shape of snow grains (Kokhanovsky and Macke (1997)):

$$A\approx 6. \qquad (4)$$

EQUATIONS TO RETRIEVE THE SNOW GRAIN SIZE

Let the snowfield without industrial pollution is observed by a satellite optical instrument. As it follows from Eq.1, the effective radius of snow grains can be retrieved using data of two spectral channel i and j in accordance with the relation

$$a_{ef}=\left[\frac{G(\theta,\theta_0)}{\alpha_j-\alpha_i}\left(\frac{R_i^{\alpha_j}}{R_j^{\alpha_i}}\right)^{\frac{1}{\alpha_j-\alpha_i}}\ln\left(\frac{R_i}{R_j}\right)\right]^2 \qquad (5)$$

where $G(\theta,\theta_0)=0.256(1+2\cos\theta)^{-1}(1+2\cos\theta_0)^{-1}$, $\alpha_i=2\pi\chi_i/\lambda_i$, χ_i is the imaginary part of snow refractive index at wavelength λ_i, which is the center of the i-th channel, the value of χ_i being known (Warren (1984)).

The retrieval of the effective grain size, using (5), does not imply any *a priory* suggestion about snow model. The price for this great advantage of this retrieval algorithm is including the extra-channel.

Recently, the algorithm to retrieve not only effective size of snow grains by the pollution amount as well, which employs the above ideas and equations, has been already developed and checked carefully.

The right choice of channels from those available for specific satellite multi-spectral instrument is also a part of the retrieval algorithm. We have shown only that these spectral channels should be chosen in the range 0.6-1.25 μm. It follows from carried out simulations the relative error of the grain size retrieval does not exceed 30% if satellite data have random error less than 1%.

ATMOSPHERIC EFFECT ON THE RETRIEVAL ALGORITHM

Up to now the algorithm to retrieve the effective snow grains size and relative soot concentration values from re-

flection coefficients at the snowfield level has been only discussed. But, satellite data is a response of an atmosphere-snow system. How to take into account the atmospheric effects? The first step is to estimate the atmospheric effect on the results of the retrieval without any changes in the retrieval algorithm.

Thus, let us suppose that satellite data (reflectance coefficients of the atmosphere-snow system) without any atmospheric correction is the input of the developed algorithm.

To simulate an atmospheric effect, we have used the conventional aerosol atmosphere model as a three layers system with different aerosols (Han W. and K.Stamnes, in press). The highest layer (10-30 km) contains the background stratospheric aerosol (sulfuric acid). Its optical thickness is 0.03 at 0.55 μm. The troposphere layer (3-10 km) has the tropospheric aerosol, which is a mixture of 70% water-soluble and 30% dust-like aerosols. The optical thickness of the tropospheric aerosol is assumed to be 0.07 at 0.55 μm. The optical properties of aerosol in the lowest layer (3 km) are described using either continental or water-soluble aerosol. The optical thickness of the lowest layer varies from 0.05 to 0.4 at 0.55 μm. Thus, the total optical thickness of aerosol atmosphere ranges from 0.15 to 0.5 at 0.55 μm.

The sub-arctic molecular atmosphere model is employed in atmosphere modeling.

In our simulation a snowfield is considered as an additional layer, which is located at the atmosphere bottom. Snow layers are modeled as assembles of chaotically oriented fractals, spheres, prolate or oblate spheroids. Their effective sizes vary from 50 to 1000 μm. The relative soot concentration ranges from 0 to 10^{-6}.

The analysis of the whole body of obtained data shows the following.

1. Variation of the aerosol type in the lowest sublayer or the snow grain shape results in the error of retrieval of the effective snow grain size and relative soot concentration less than 10%.
2. The error of retrieval of the effective snow grain size due to atmosphere present is less than 40% and the error of retrieval of the relative soot concentration due to atmosphere present is less than 100%, if snow is "no black". The snow is "black", if the spherical snowfield albedo is less than 0.5 at 0.68 μm and less than 0.3 at 0.865 μm. It means that effective snow grain size and relative soot concentration are supposed not be very large at the same time.

So, the conclusion is: our Retrieval Algorithm could be used even **without atmospheric correction** excluding cases of very turbid atmosphere (the optical thickness of an atmosphere $\tau_{atm}(\lambda = 0.55\,\mu m) > 0.3$) and old and heavily polluted snow (effective snow grain size>500 μm and relative soot concentration $>5.10^{-7}$).

However, comparison of retrieval accuracy in two cases, when the retrieval algorithm input is:

- an immediate spectral response of snow layer or
- spectral response of atmosphere-snow layer,

shows that atmospheric correction can noticeably increase accuracy of the retrieval. The development of atmospheric correction method within developed algorithm is our current problem.

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REFERENCES

Bohren C.F., 1974: Theory of the Optical properties of Snow. *J.Geoph.Res.* **79**, 4527-4535.

Greenfell J.C., S.G. Warren and P.C. Mallen, 1994: Reflection of solar radiation by the Antarctic snow surface at ultraviolet, visible, and near-infrared wavelength. *J.Geoph.Res.*, **99**, 18669-18684.

Han W. and K.Stamnes, in press: Remote sensing of Surface and Cloud Properties in the Arctic from AVHRR Measurements, J.Appl.Meteorol.

Ivanov A.P. et al, 1988, *Light scattering in close packed media,* Nauka i Texnika, Minsk, 191 p.

Kokhanovsky A.A and A. Macke , 1997: Integral light-scattering and absorption characteristics of large, nonspherical particles. *Appl.Opt.*, **37**, 8785 - 8790.

Stamnes K., B. Chen and W. Han, 1997: Remote sensing of cloud and aerosol properties in the arctic region from GLI measurments on board ADEOS-II. *in press.*

Warren S.G., 1984: Optical constant of ice from the ultraviolet to the microwave. *Appl.Opt.*, **23**, 1206-1233.

Wiscombe W.J. and S.J.Warren, 1980: A Model for the Spectral Albedo of Snow. I. Pure Snow. *J.Atmos.Sci.*, **37**, 2712-2733.

___________ and S.J.Warren, 1980a: A Model for the Spectral Albedo of Snow. II. Snow Containing Atmospheric Aerosols. *J. Atmos. Sci.*, **37**, 2734-2745.

Zege, E.P, A.P.Ivanov and I.L.Katsev, 1991: *Image Transfer through a Scattering Medium,* Springer-Verlag, 352 pp.

_______. A.A. Kokhanovsky, 1997: Aproximated fomula for snowfield albedo. *Izv. RAN, Ser. Fiz. Atm. Okean*, **33**, 719-720.